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A note on infragravity waves in the Arctic Ocean

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Abstract. Long-period (20–50 s) surface-gravity waves were recorded in the eastern Arctic (83°N, 10°E) using tiltmeters deployed on the ice and an acoustical current meter deployed a few meters below the ice during the spring of 1989. The waves caused ice tilt of a few microradians and horizontal velocity of the water just beneath the ice of order 10^{-4} m s^{-1}. The unprecedented sensitivity of the velocity measurements is the result of a path-averaging technique based on reciprocal transmission along 200-m horizontal paths. The measurements indicate that infragravity waves and low-frequency swell penetrate the Arctic Ocean through the Fram Strait and that they are freely propagating and not tied to groups of short waves. It is suggested that these waves could provide a method to measure ice thickness distribution in the Arctic Ocean.

Introduction

Recently, *Webb et al.* [1991] reported pressure measurements from instruments sited on the deep seafloor indicating that infragravity waves (surface-gravity waves with periods longer than the swell and wind-driven waves) are a ubiquitous feature of the Atlantic and Pacific Oceans. They found that these waves are freely propagating and not tied to groups of short waves. Here we report the observation of infragravity waves in the eastern Arctic using tiltmeters deployed on the ice and a path-averaging acoustical current meter. These measurements are of interest, both because they support the conclusion reached by *Webb et al.* [1991] regarding the ubiquitous presence of freely propagating infragravity waves in the deep ocean and because they demonstrate the sensitivity of the path-averaging acoustical current meter.

Infragravity waves cannot be directly generated by the wind in the deep ocean where their phase speed exceeds that of the strongest winds. The major portion of infragravity wave energy is refractively trapped in coastal regions where these waves are believed to be important for many nearshore processes and where they have been extensively studied [*Elgar et al.*, 1992]. Interest in measuring infragravity waves in the deep ocean was initiated by *Webb et al.* [1991], who suggested that these waves can provide a new method to study the shallow elastic structure of the oceanic crust. We are intrigued by the possible use of infragravity waves in the Arctic Ocean to remotely monitor ice thickness distribution.

Surface-gravity waves from the open ocean propagate into the Arctic as flexural-gravity waves [*Wadhams*, 1986]. Scattering at the ice edge attenuates swell and high-frequency surface waves. In addition, leads, irregularities in the ice, and plastic creep dissipate swell energy within the pack ice, attenuating short wavelengths more strongly than long ones. In effect, the ice acts as a low-pass filter whose wavenumber cutoff decreases with increasing distance from the open ocean. For this reason, infragravity waves that manage to reach the central pack are almost certainly freely propagating and not tied to groups of short waves.

The surface waves reported here cause ice tilt of a few microradians and horizontal velocity of the water just beneath the ice of order 10^{-4} m s^{-1} at periods of 20-50 s. Such a small signal would be masked by turbulent fluctuations in the near-surface boundary layer were it not for the integrating nature of the measurements. Tilt measurements are inherently integrating owing to the properties of the ice cover. The current measurements were obtained using acoustical reciprocal transmission along a 200-m horizontal path. This configuration at-

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tenuates the shorter wavelengths due to turbulence and allows the detection of infragravity waves.

Measurements

The measurements were obtained at the oceanography ice camp of the Coordinated Eastern Arctic Experiment (CEAREX) established 300 km northwest of Spitsbergen (83°N, 10°E) during the spring of 1989. Three electrolytic bubble level tiltmeters were frozen to the surface of the ice in a triangular array [*Czipott et al.*, 1991]. One axis of each meter pointed north and the other, east. Horizontal water velocity near the surface, averaged over 200-m horizontal paths, was measured with a triangular array of acoustic transducers suspended 20 m below the ice [*Menemenlis and Farmer*, 1992]. The acoustic array was centered about 337 m to the south of the tiltmeter array.

Spectra of north-south and east-west ice tilt for a 5-hour period starting at 1300 UTC on April 10, 1989, are displayed on Figure 1. The spectra exhibit a broad peak near the 30-s period. Cross-spectral analysis of the two orthogonal axes indicates that infragravity waves (~50 s) come mainly from the western Fram Strait (~220°T) as they can travel long distances with little attenuation, while long-period swell (~20 s) comes predominantly from the eastern Fram Strait (~195°T) where it passes through less ice [*Czipott and Podney*, 1989].

The corresponding path-averaged horizontal velocity spectrum for an acoustic path oriented 223°T at the 20-m depth is drawn on Figure 2 (solid line). We will show that the spectral peak centered at 0.03 Hz is due to infragravity waves and is consistent with the tilt measurements. The 0.08-Hz spectral peak results from relative mooring motion during each reciprocal transmission. *Menemenlis and Farmer* [1992] showed that this contribution to the spectrum is significant only near the moorings' mechanical resonance and is not responsible for the 0.03-Hz spectral peak.

The mean ice thickness and depth of the water column at the ice camp during the measurements were 2.4 and 2400 m, respectively. For a deep ocean, surface-gravity waves with 20-to-50-s periods correspond to 600-to-4000-m wavelengths. At these wavelengths the ice behaves as a flaccid membrane [*Czipott et al.*, 1991], and the dispersion relation of freely propagating surface-gravity waves [*Pond and Pickard*, 1983, Section 12.3],

$$\omega^2 = gk, \tag{1}$$

is appropriate, where ω is the angular frequency, g is the acceleration due to gravity and k is the wave number. Consider a wave with surface displacement $\eta(\mathbf{x}, t) = \eta_o \cos(\mathbf{k} \cdot \mathbf{x} - \omega t)$ and near-surface horizontal velocity $u_h(\mathbf{x}, t) = \omega \eta(\mathbf{x}, t)$; η_o is the wave amplitude, $\mathbf{k}$ is the vector wavenumber, $\mathbf{x}$ is position and t is time. Ice tilt τ is then related to surface displacement η and horizontal velocity u_h by

$$\tan \tau = \nabla_h \eta = \frac{1}{\omega} \nabla_h u_h, \tag{2}$$

where ∇_h is the horizontal gradient operator. For small

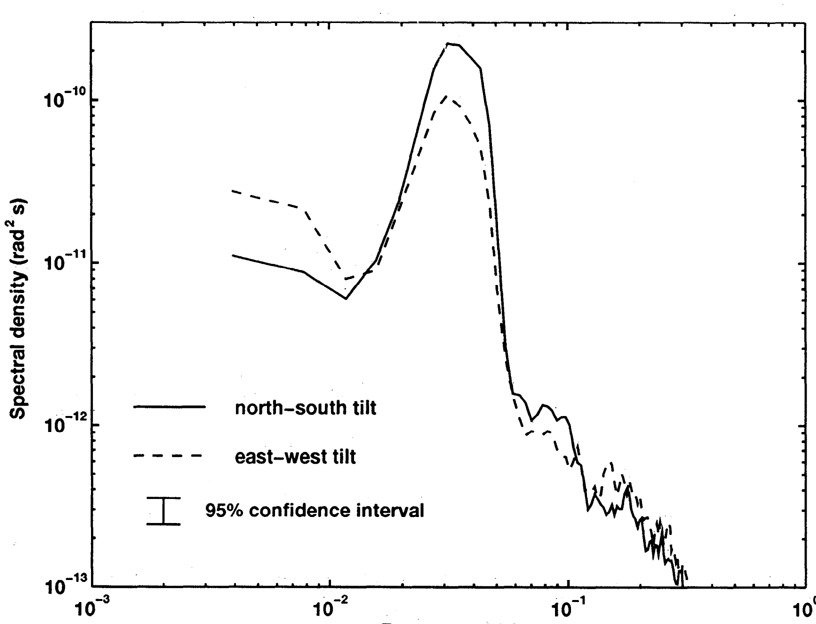

Figure 1. Spectra of north-south (solid line) and east-west (dashed line) ice tilt in the eastern Arctic (83°N, 10°E), for a 5-hour period starting at 1300 UTC on April 10, 1989. Cross-spectral analysis of the two orthogonal axes indicates that infragravity waves (~50-s period) come mainly from the western Fram Strait (~220°T) as they can travel long distances with little attenuation, while long period swell (~20 s) comes predominantly from the eastern Fram Strait (~195°T) where it passes through less ice.

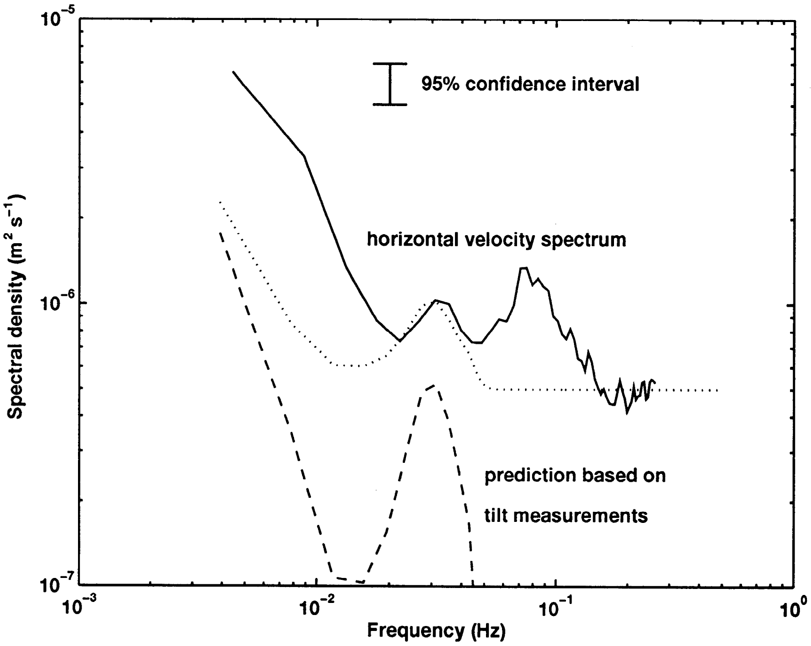

Figure 2. Line-averaged horizontal velocity spectrum (solid line) along a 200-m measuring baseline oriented 223°T, located 20 m below the ice cover and 337 m to the south of the tiltmeter array, for the same time period as in Figure 1. The dashed line is a prediction based on the tilt measurements of horizontal velocity fluctuations caused by surface-gravity waves. This comparison suggests that the 0.03-Hz spectral peak can be attributed to freely propagating infragravity waves originating in the western Fram Strait. To explain the 50% discrepancy between the tilt and the velocity measurements, white noise was added to the model at a level consistent with that observed at high frequency (dotted line).

surface displacements relative to the wavelength, yields

$$\tau_o = k\eta_o = \frac{k}{\omega} u_{ho}, \tag{3}$$

where τ_o, η_o, and u_{ho} are the amplitudes of ice tilt, surface displacement, and horizontal velocity, respectively.

The path-averaging current meter measures the mean ice–relative velocity along each 200-m acoustic path with a response function given by [*Menemenlis and Farmer*, 1992]

$$u_m = \frac{\sin(\mathbf{k} \cdot \mathbf{l}/2)}{kl/2} u_h, \tag{4}$$

where $\mathbf{l}$ is the path vector and l is the range. This equation is valid for long disturbances such as infragravity waves which remain coherent along the acoustic path. A different expression must be used for short incoherent disturbances such as turbulent velocity fluctuations [*Menemenlis*, 1994]. Neglecting the Doppler shift due to the mean ice–relative water velocity and assuming that the acoustic path is aligned with the incoming waves, (1), (3), and (4) relate ice tilt amplitude τ_o to the amplitude of the line-averaged horizontal velocity fluctuations u_{mo},

$$\frac{u_{mo}}{\tau_o} = \frac{2g^2}{l\omega^3} \sin\left(\frac{l\omega^2}{2g}\right). \tag{5}$$

The spectrum of near-surface horizontal velocity predicted from ice tilt measurements is drawn as a dashed line on Figure 2. Both the predicted and the measured spectra exhibit a peak at 0.03 Hz, but there is a 50% discrepancy in the spectral density. We attribute this discrepancy to uncorrelated additive noise in the velocity measurements and to the calibration uncertainty of the tilt measurements. Specifically, the discrepancy could be explained by white noise in the velocity measurements at a level of 5×10^{-7} m^2 s^{-1}, consistent with that observed at high frequency (dotted line on Figure 2). In addition, the tiltmeters were found to have anomalously high temperature coefficients during the experiment. *Czipott and Podney* [1989] reported site-to-site amplitude variations by a factor of 6 between concurrent tilt measurements at the ice camp. The site-to-site variation was reduced by applying correction factors obtained during cold-chamber tests following the experiment, but a residual factor of 2 discrepancy could not be eliminated [*Czipott*, 1991].

Discussion

The initial motivation for comparing ice tilt and horizontal velocity measurements at infragravity wave periods was to test the sensitivity of the acoustical current meter. During the observations the mean ice–relative velocity at the 20-m depth was 0.06 m s^{-1} in a southward direction. At this rate, 2-m-sized turbulent eddies with 10^{-3} m s^{-1} characteristic velocity are responsible for 30-s-period velocity fluctuations relative to the ice. The observed infragravity waves have characteristic velocity of order 10^{-4} m s^{-1} and cannot be detected with point measurements of current. Equation (4) states that the acoustical current meter is a spatial filter with 450-m cutoff wavelength which is short compared to the wavelength of infragravity waves, but long compared to turbulence scales. Path averaging makes detection of the weak signal associated with infragravity waves possible.

Within experimental error the observations presented here are consistent with the assumptions of thin ice, deep ocean, and freely propagating infragravity waves. The measurements indicate an rms surface displacement of a few tenths of a millimeter. This value is consistent with previously reported gravity measurements on the Arctic Ocean [*Hunkins*, 1962; *LeSchack and Haubrich*, 1964] and the Beaufort Sea [*Crary et al.*, 1952] and to values inferred from seafloor pressure measurements in the Beaufort Sea [*Webb and Schultz*, 1992] and the eastern Atlantic [*Webb et al.*, 1991]. Our measurements also indicate that infragravity waves and low-frequency swell penetrate the Arctic Ocean through the Fram Strait. We suggest that these waves could be used for remote monitoring of ice thickness distribution.

Ice thickness distribution is needed for climatological and forecast modeling [*Rothrock*, 1986], but no satisfactory remote sensing techniques are available for operational use at present [*Carsey and Zwally*, 1986; *Shuchman and Onstott*, 1990]. The loss mechanisms for the slow decay of flexural-gravity waves depend on ice thickness, type, and concentration (including floe size distribution and location of leads) and on the oscillation period of the waves [*Wadhams*, 1986]. Assuming the bulk of infraswell energy is not generated locally, then the slow decay of the waves as they propagate through the ice-covered ocean provides an integral measure of sea ice properties.

To make the discussion more quantitative, consider the dispersion relation described by *Squire and Allan* [1980] for periodic waves with attenuation due to the creep properties of sea ice,

$$\frac{h^3}{6}(b_0\omega + ib_1)k^5 + (\omega - \frac{a_2}{\omega} + ia_1)(\rho g - \rho_i h\omega^2)k$$
$$-\rho(\omega - \frac{a_2}{\omega} + ia_1)\omega^2 = 0, \qquad (6)$$

where h is the ice thickness, ρ and ρ_i are the densities of seawater and sea ice, respectively, and a_i and b_i are material constants related to the viscoelastic properties of the ice. It is readily verified that for $h = 0$, (6) reduces to (1), the deepwater, open-sea form. We used the values $a_1 = 1.4 \times 10^{-2}$ s^{-1}, $a_2 = 0.07 \times 10^{-4}$ s^{-2}, $b_0 = 1.2 \times 10^9$ N m^{-2}, and $b_1 = 1.02 \times 10^7$ N m^{-2} s^{-1} suggested by *Tabata* [1958], set $\rho = 1028$ kg m^{-3} and $\rho_i = 900$ kg m^{-3}, and solved (6) for the attenuation coefficient as a function of frequency and ice thickness. The north-south tilt spectrum of Figure 1 is redrawn on Figure 3, along with the projected attenuation due

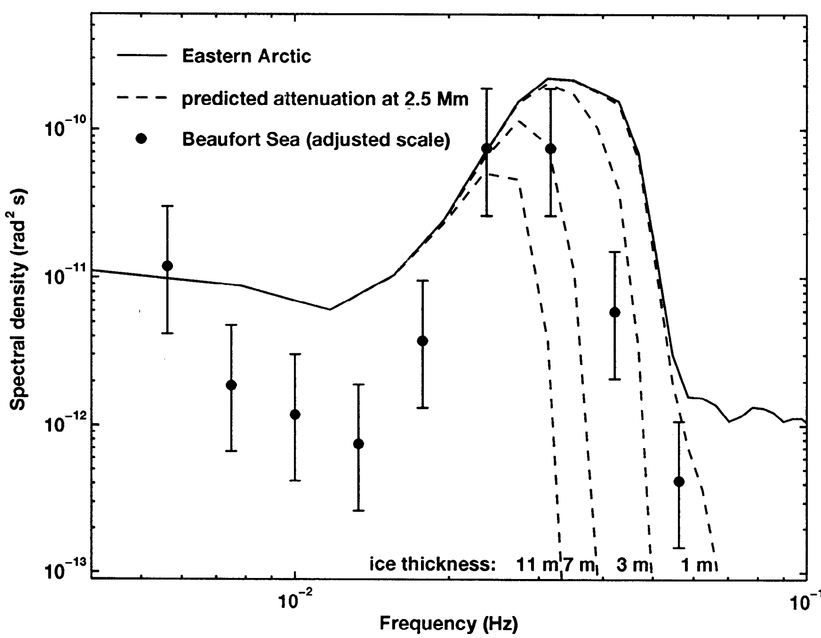

Figure 3. Portion of the north-south tilt spectrum of Figure 1 (solid line), with projected attenuation due to plastic creep at a range of 2.5 Mm (dashed line) for uniform, intervening ice. The cutoff frequency is seen to be a sensitive function of ice thickness. Also shown are north-south tilt measurements obtained by *Czipott and Podney* [1985] in the Beaufort Sea (74°N, 145°W) on the night of March 18–19, 1985 (solid circles with error bars). The lower cutoff frequency of the Beaufort spectrum may be an indication that a major portion of the infraswell energy is not generated locally but that it has propagated hundreds of kilometers in the ice-covered ocean. It is suggested that these types of measurements obtained simultaneously at a few key locations in the Arctic Ocean could be used to remotely monitor ice thickness distribution.

to plastic creep at a range of 2.5 Mm. At these ranges the attenuation of infragravity swell by the ice cover is clearly a sensitive function of ice thickness.

Wadhams [1973] describes a different model, based on non-Newtonian ice creep, where the attenuation coefficient is a function of wave amplitude. He showed this model to be compatible with observations available at the time. However, this model predicts essentially no attenuation for waves with periods longer than 13 s and amplitudes smaller than 1 mm, and therefore it is incompatible with the cutoff frequency of the present observations. We postulate that Newtonian damping is a better model for the attenuation of surface waves by the ice in the central Arctic.

Also shown on Figure 3 is the spectral density of tilt measurements from the Arctic Internal Wave Experiment (AIWEX) [*Czipott and Podney*, 1985]. The AIWEX measurements were obtained during the spring of 1985 on the Beaufort Sea, 4 years before and 2.5 Mm away from the site of the CEAREX ice camp. Nevertheless, the qualitative agreement between the spectra suggests that the low-frequency end of the infraswell spectrum may have a consistent form in the Arctic Ocean. The erosion of the AIWEX spectrum at the higher frequencies also suggests that the bulk of infraswell energy was not generated locally but that it may have propagated for hundreds of kilometers before reaching the site of the experiment.

Evidently, an attenuation model based on a uniform, continuous ice sheet is but a first-order approximation. As discussed by *Wadhams* [1986], the model needs to be augmented by considering the effects of ice inhomogeneities, leads, polynyas, pressure ridges, hummocks, as well as the local generation of surface-gravity waves. Nevertheless, we envisage that simultaneous directional measurements of surface-gravity waves at a few key locations on the Arctic Ocean could be combined with spaceborne measurements of ice type and concentration and with improved propagation models, in order to monitor path-averaged ice thickness.

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